Chapter 1

Mesoscale Modeling

1.1 A brief historical perspective

On Earth, the first comprehensive mesoscale studies were often motivated by concerns about meteorological hazards at regional scales: in the 1970's, numerical models were employed to complement earlier investigations based on theoretical considerations [e.g. tropical hurricanes: Yanai, 1961, Anthes et al., 1971] and data analysis [e.g. flow impinging on a topographical obstacle: Mahrer and Pielke, 1977, Lilly and Kennedy, 1973]. Those early terrestrial efforts yielded idealized two-dimensional modeling studies of local atmospheric disturbances in the Martian atmosphere, especially ones associated with topography: slope winds [Ye et al., 1990, Savijärvi and Siili, 1993] and lee waves [Pickersgill and Hunt, 1981, Tobie et al., 2003]. In the 1980's and beyond, a more unified approach was adopted in terrestrial studies. The modeling efforts were pursued to build platforms capable of reproducing the mesoscale variability in any region of the world at various horizontal scales [Pielke et al., 1992, Dudhia, 1993]. Those efforts yield the versatile three-dimensional mesoscale models used nowadays [Skamarock and Klemp, 2008] in operational weather prediction, meteorogical research and planetary science studies. In the 2000's, both this availability of flexible regional-climate models on Earth and an harvest of new observations of Mars, some of which related to phenomena left unresolved by global circulation models [GCMs], motivated the development of dedicated three-dimensional mesoscale models for the Martian atmosphere [Rafkin et al., 2001, Tyler et al., 2002, Toigo and Richardson, 2002, Siili et al., 2006, Wing and Austin, 2006, Richardson et al., 2007, Spiga and Forget, 2009].

Those efforts gave birth to powerful simulators of the Martian atmospheric circulations at the mesoscale (100s of km – 1 km, this chapter) and the microscale (1 km – 100s of m, see chapter **??**). Meteorological models aiming at resolving mesoscale phenomena share similar structure and design as GCMs described in chapter **??**. They comprise of two main modules:

1. the *dynamical core* integrates primitive equations for the atmospheric fluid, i.e. Navier-Stokes equations projected on the rotating frame using spherical coordinates.

2. the key diabatic forcing in those equations are computed through so-called *physical parameterizations* for radiative transfer, surface-atmosphere heat

exchanges, latent heat release, friction; as well as unresolved dynamical phenomena under the chosen grid spacing, e.g. boundary layer mixing.

Despite this common framework, crucial differences can be reported between numerical models suitable for large-scale circulation and for mesoscale phenomena.

1.2 (Non-)Hydrostaticity

While GCMs assume hydrostatic equilibrium in their built-in primitive equations given the involved low vertical velocities at synoptic scale, this could not be the case for dynamical cores in mesoscale models, which have the ambition to resolve atmospheric phenomena at finer scales. The hydrostatic equilibrium consists in retaining gravity only as being responsible for vertical stratification of pressure:

$$\frac{\partial p}{\partial z} = -\rho g \tag{1.1}$$

with altitude z, pressure p, acceleration of gravity g and density ρ. An alternative formulation for hydrostaticity is to define it by integrating the above equation between two atmospheric levels z_1 and z_2

$$p_1 - p_2 = -\int_{z_1}^{z_2} \rho\, g\, dz \tag{1.2}$$

where pressure arises naturally as an equivalent for atmospheric mass.

Hydrostatic equilibrium is only strictly true for a static atmosphere, devoid of any horizontal pressure gradients, which is not the case of a real atmosphere, as can be noticed from e.g. geostrophic equilibrium. Notwithstanding this, dimensional analysis and observations show that, for most meteorological phenomena developping at spatial scales larger than $\sim$ 10 km, gravity force and vertical pressure gradient remain the two most prominent terms in the vertical component of primitive atmospheric equations. Hydrostatic equilibrium stands as a correct approximation as long as acceleration of vertical wind $\frac{dw}{dt}$ is negligible compared to acceleration of gravity g. To illustrate this, Janjic et al. [2001] proposed a simple equation based on the distinction between barometric pressure p_s at the surface and hydrostatic (mass-based) pressure π_s

$$p_s = \pi_s + \int_0^1 \epsilon \pi(\sigma')\, d\sigma' \quad \text{with} \quad \epsilon = \frac{1}{g}\frac{dw}{dt} \tag{1.3}$$

where σ' denotes a convenient vertical coordinate which definition is not essential here. As long as vertical accelerations in the atmosphere are low or inhibited, e.g. when atmospheric stability is particularly high, hydrostaticity can be assumed.

While hydrostaticity is still fairly common amongst regional-scale phenomena, mesoscale dynamics possibly involve significant departures from hydrostatic equilibrium: convective updrafts, gravity waves, ... Hence the equation for vertical motions is often implemented in its complete version in mesoscale models. The importance of non-hydrostatic effects can be further illustrated through considering mesoscale gravity waves, associated to the buoyancy restoring force:

the dispersion relation between wave frequency ω and spatial wavenumber (k, l, m) writes

$$\omega^2 = f^2 + N^2 \, \frac{k^2 + l^2}{m^2} \tag{1.4}$$

under hydrostatic assumption, but writes

$$\omega^2 = f^2 \, \frac{m^2}{k^2 + l^2 + m^2} + N^2 \, \frac{k^2 + l^2}{k^2 + l^2 + m^2} \tag{1.5}$$

with all non-hydrostatic contributions [see Fritts and Alexander, 2003, f is Coriolis parameter and N is atmospheric stability]. In other words, hydrostatic integration left an important part of the gravity wave spectrum unresolved.

1.3 Initial and boundary conditions

While GCMs integrate the geophysical fluid equations on the whole sphere, the vast majority of mesoscale modeling for the terrestrial and Martian atmosphere make use of limited-area strategy. Computations are carried out in a specific region of interest on the planet. An adapted map projection is defined for the chosen region of computation: e.g. stereographic projections are used for polar regions Tyler and Barnes [2005], which ensure mesoscale simulations are devoid of any pole singularity, an usual drawback of the GCMs that requires the use of additional filtering. Note that another approach to resolve mesoscale phenomena through atmospheric modeling is to run GCMs at high resolution or use adaptable-grid zooming capabilities [Moudden and McConnell, 2005, **?**]. This approach is not prominent thus far, as only resolutions of ~ 40 km have been achieved, which is much coarser than limited-area mesoscale models. This might change in the future, owing to progress in computational efficiency.

1.3.1 Horizontal dimension

Horizontal boundary conditions for the main meteorological fields (horizontal winds, temperature, tracers) have to be provided during the simulations, in addition to an atmospheric starting state. Idealized simulations usually require the use of periodic, symmetric or open boundary conditions, whereas real-case simulations need specified climatologies at the boundaries. In Martian mesoscale modeling, the specified boundary conditions and the atmospheric starting state are derived from previously performed GCM simulations which have reached equilibrium, typically after ~ 10 simulated years. A relaxation zone of a few grid points width is implemented at the boundaries of the mesoscale domain to enable both the influence of the large-scale fields on the limited area, and the development of the specific mesoscale circulation inside the domain. GCM fields are usually updated every one or two Martian hour to constrain the mesoscale model at the domain boundaries: this higher frequency in large-scale updates than in the terrestrial case (closer to 6 hours) is deemed necessary by strong thermal tide perturbations on Mars, as first noticed by Tyler et al. [2002].

In terrestrial mesoscale modeling meteorology, Dimitrijevic and Laprise [2005] showed, by the so-called "Big Brother" approach, that the single-domain approach yields unbiased results when the boundary forcing involves a minimum of $\sim 8-10$ GCM grid points [possibly lower in situations of complex topography, as

shown in Antic et al., 2006]. Hence the single-domain approach is only suitable for mesoscale simulations of horizontal resolution of $dx \sim 10$s km; to reach finer resolution of few kilometers, nested domains must be used as first introduced in numerical studies of terrestrial fronts [Hill, 1968, Harrison and Elsberry, 1972]. The nested simulations feature two kinds of domains where the meteorological fields are computed: the "parent" domain, with a large geographical extent, a coarse grid resolution, and specified boundary conditions, and the "nested" domains, centered in a particular zone of interest, with a finer grid resolution, and boundary conditions provided by its parent domain. Using parent domain sufficiently extended over the planet is necessary to correctly represent thermal tides and planetary waves (e.g. Kelvin, Rossby waves). Spiga and Forget [2009] showed that wraping up a mesoscale model all around the planet in the longitudinal dimension allowed for consistent modeling of large-scale atmospheric waves (such simulations are sometimes named "tropical channel" simulations).

1.3.2 Vertical dimension

Vertical boundaries (bottom and top of the model) have to be treated as well in mesoscale simulations. At the top of the domain, free relaxation condition to zero vertical velocity is usually imposed. Gravity wave absorbing layers can be defined as well to dampen those disturbances which amplitude is larger at higher altitudes. Mesoscale simulations are usually carried out with lower model top than GCMs, which could adversely affect results near the model top which might not accurately represent the actual flow as discussed by Toigo and Richardson [2002] and Spiga and Forget [2009]; for this reason, mesoscale simulations have adressed Martian meteorological phenomena below 40 km altitudes. Surface boundary is usually treated through physical parameterizations, while topography is taken into account in the dynamical core through calculating a surface geopotential. Surface static data intended for the mesoscale domain are extracted from maps derived from recent spacecraft measurements: 64 pixel-per-degree (ppd) MOLA topography [*Smith et al.*, 2001], 8 ppd MGS/Thermal Emission Spectrometer (TES) albedo [*Christensen et al.*, 2001], 20 ppd TES thermal inertia [*Putzig and Mellon*, 2007].

In the process of initialization and definition of boundary conditions, the vertical interpolation of GCM meteorological fields to the terrain-following mesoscale levels must be treated with caution. While deriving the near-surface meteorological fields from GCM inputs, one may address the problem of underlying topographical structures at fine mesoscale horizontal resolution, e.g., a deep crater that is not resolved in the coarse GCM case. A crude extrapolation of the near-surface GCM fields to the mesoscale levels is usually acceptable for terrestrial applications. On Mars, owing to the low density and heat capacity of the Martian atmosphere, the surface temperature is to first order controlled by radiative equilibrium, and thus it is left relatively unaffected by variations of topography [e.g., Nayvelt et al., 1997]. A practical consequence, which renders an extrapolation strategy particularly wrong on Mars, is that the near-surface temperature and wind fields vary much more with the distance from the surface than with the absolute altitude above the areoid (or equivalently with the pressure level). Initial tests carried out with the extrapolation strategy showed that differences between temperatures at the boundaries and temperatures computed within the mesoscale domain close to these boundaries often reach $20-30$ K

near the surface.

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